Volume loss of the Greenland ice sheet revealed by SARAL/AltiKa repeat passes radar altimetry

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The Greenland ice sheet (GrIS) is one of the drivers of global sea level rise and plays a crucial role in understanding the global climate changes. Here, we have estimated and analysed the decadal (between 2013–2016 and 2003–2005) and annual (2014–2015, 2015–2016) volume discharge of ice from the entire GrIS. The 40 Hz Geophysical Data Record product of the unique Ka band (AltiKa) radar altimeter were utilised to derive the elevation, elevation changes and volume changes over the GrIS. To test the first-level accuracy of the result, AltiKa and NASA's ice, cloud and land elevation satellite digital elevation model (ICESat DEM)-derived elevation were compared, which yielded a correlation value of 0.95. Thereafter, decadal volume changes obtained over the entire GrIS, from the differencing of the AltiKa and ICESat DEM elevation revealed a decreasing rate of 247 km³/year. Moreover, basin-wise analysis indicated the maximum decrease in elevation of basin located in the north and north-west region of GrIS. Annual changes obtained by differencing the AltiKa cycle of the same month (so that the surface condition will remain same) between the two consecutive years, specifically during 2014–2015 and 2015–2016 over the entire GrIS contributed volume loss of 187 and 210 km³, respectively, indicating an enhanced decrease for a later period.

Keywords. Greenland ice sheet; elevation change; volume change; SARAL/AltiKa; radar altimeter.

1. Introduction

The Greenland ice sheet (GrIS) and its response to global climate changes are one of the major concerns to scientific community. Indeed, GrIS is the largest exclusive source of sea-level rise and will remain so for decades as per the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (Van den Broeke *et al.* 2017). This invokes an urgent need for its change estimation. The GrIS being remote, vast and hostile, remote sensing has played a key role in observing and predicting the changes occurring over it. Three main approaches utilised to assess the polar ice sheet changes are the

input–output method, gravimetry and altimetry. First, one is a ground-based method while the other two are remote sensing-based methods. Recently, Wang *et al.* (2018) used gravity recovery and climate experiment gravity field and surface mass balance data (RACMO2.3) of duration 2003–2015 to study GrIS. The third approach is altimetry, so far various laser and radar altimeters were widely utilised in mapping the topography (DiMarzio *et al.* 2007; Helm *et al.* 2014) and estimating the discharge of GrIS (Krabill *et al.* 2004; Li and Davis 2008; Helm *et al.* 2014). All of these studies show a high thinning rate, near the coastal region and stable or slight thickening at the interior of the

GrIS. Furthermore, the thinning rate was measured to be higher below 2000 m elevation, as compared to the elevation above 2000 m (Krabill *et al.* 2000). A drastic thinning rate was observed over the fastest moving Jakobshavn glacier, located near the coast in the south-western region of the GrIS (Rignot and Kanagaratnam 2006). Thereby, it adds high amount of ice mass discharge into the ocean and acting as a potential driver to sea-level rise.

However, altimeter data require retracking and slope correction to use it for ice-sheet studies (Bamber 1994). Due to surface irregularities, tracking gate fails to track the midpoint of the leading edge, which leads to an incorrect measurement of the range of the satellite and henceforth incorrect elevation measurement. Therefore, ‘retracking’ is required to track the midpoint of the leading edge accurately (Bind-schadler *et al.* 1989). Secondly, the slope of the terrain displaces the range measurement from the original nadir point introducing the ‘slope induced error’ of a few kilometres for a satellite at an altitude of 800 km (Hurkmans *et al.* 2012). To mitigate the effect of this error, slope correction is required.

Apart from the above-mentioned error, high correlation observed between backscatter change and elevation change over some regions of interest (ROI) of the Antarctic ice sheet led to an error in the measurement of elevation changes (Davis and Ferguson 2004; Qiang *et al.* 2015). However, Brenner *et al.* (2000) concluded that corrections to dh/dt for variations in backscatter are not appropriate, at least for Greenland.

Most of the altimeter-based studies mentioned above were carried out using either Ku band radar altimeters or laser altimeters while very few studies (over GrIS) have been reported with Ka band (Chander *et al.* 2015; Yang *et al.* 2018). Ka band frequency being higher than Ku penetrates less into the surface giving rise to near subsurface volume echo and reduces ionosphere correction to be applied to the height measurement. Sharp antenna aperture will give a stable height measurement over time and a higher pulse repetition frequency gives along the track spacing of 175 m for a 40 Hz dataset better than 370 m as in the case of the 20 Hz dataset of an earlier altimeter mission like Envisat (Remy *et al.* 2014).

At present, SARAL/AltiKa (launched on 25 February 2013) is the only altimeter with Ka (35.75 GHz) band frequency. In the present study,

we have utilised SARAL/AltiKa Exact Repeat Mission (ERM) dataset and NASA’s Ice, Cloud and land Elevation Satellite Digital Elevation Model (ICESat DEM) dataset to quantify decadal and annual elevation and volume changes occurring over the GrIS. Along with the Airborne Topographic Mapper (ATM) dataset (<https://nsidc.org/icebridge/portal/map>) was also utilised for the validation of AltiKa-derived elevation changes, which is described in section 3.1.

2. Data and methodology

We have utilised 40 Hz Geophysical Data Record (GDR) of SARAL/AltiKa dataset (ftp://avisoftp.cnes.fr/AVISO/pub/saral/gdr_t) of the ERM phase. ICESat DEM (DiMarzio *et al.* 2007) of resolution 1000 m was used, to incorporate slope (less than or equal to 1°) correction in elevation measurements of SARAL/AltiKa. DiMarzio *et al.* (2007) validated the ICESat DEM dataset with ATM data over Greenland and it has been observed that the mean difference varies with slope: for regions with less than 0.1° slopes, the mean difference is -32 ± 43 cm; for regions with 0.1 – 1.0° slopes, the mean difference is -66 ± 61 cm. Details of the dataset used for the entire study are mentioned in table 1.

2.1 Method for deriving the elevation

The 40 Hz GDR range and altitude measurements of the SARAL/AltiKa dataset were used to derive elevation (w.r.t. WGS84) over the GrIS. The range parameter was derived from the two-way travel of the transmitted microwave pulse. Atmospheric interaction and tide effect increases the time delay, thereby lengthening the range than the actual value. Hence, the range was corrected for various atmospheric and tidal delays:

$$\begin{aligned} \text{Elevation} = & \text{ALT} - \text{range} - \text{DTC} - \text{WTC} - \text{IC} \\ & - \text{SETC} - \text{PTC}, \end{aligned} \quad (1)$$

where ALT is the altitude, DTC is the dry troposphere correction, WTC is the wet troposphere correction, IC is the ionosphere correction, SETC is the solid earth tide correction and PTC is the pole tide correction.

We have used an Ice-2 retracker-derived range in the present study as Ice-2 retracker is considered

Table 1. Summary of the dataset used in the present study.

Dataset	Phase of the mission	Cycle	Start date	End date
SARAL/AltiKa 40 Hz GDR dataset	ERM	01	14 March 2013	18 April 2013
		11	27 February 2014	03 April 2014
		12	03 April 2014	08 May 2014
		22	19 March 2015	23 April 2015
		23	23 April 2015	28 May 2015
		32	03 March 2016	07 April 2016
		01–35	14 March 2013	04 July 2016
ICESat DEM dataset	–	–	01 February 2003	30 June 2005

appropriate for the ice sheet study (Legresy *et al.* 2005; Remy *et al.* 2014; Chander *et al.* 2015).

Later on, these elevations were corrected for the slope-induced error, using the direct method (Brenner *et al.* 1983). This correction is proportional to the square of the slope value and the altitude of the satellite. A detail regarding the slope correction is mentioned in Suryawanshi *et al.* (2019). These slope uncorrected and corrected elevation datasets were compared with ICESat's DEM-derived elevation to check the first level accuracy of the dataset.

After slope correction, the elevation map for individual cycles (mentioned in table 1) and the average elevation map (here onwards referred to as the AltiKa35 elevation map) obtained using all 35 repeat cycles (duration: 14 March 2013–4 July 2016) of SARAL/AltiKa were generated for further analysis. AltiKa being a radar altimeter will penetrate deeper into the surface in comparison with ICESat, which is a laser altimeter. Hence, before generating the elevation difference between them, it becomes mandatory to apply correction due to penetration in the AltiKa35 elevation. To measure this bias, the AltiKa elevation derived from cycle 12 (April–May 2014) was interpolated at ATM measured elevations of 05 April 2014 over a part of GrIS and bias was calculated. Indeed, we should compare the ICESat and AltiKa datasets of the same location and same time to determine bias due to the Ka band. However, different operational time periods of these two satellites prohibited this comparison. Hence, a comparison between ATM and AltiKa was carried out, as ATM is also a laser altimeter and has a spatial and temporal overlap with the AltiKa dataset.

Furthermore, ROI-1 (76–77°N and 34–35°W) and ROI-2 (69–70°N and 47–49°W) with average slope values of 0.19° and 0.60° were selected for the validation of our elevation change results. ATM dataset

measured elevation change between April–May 2014 and April–May 2015 was compared with AltiKa-derived slope corrected and outliers removed gridded elevation difference of cycle 23 (April–May 2015) and cycle 12 (April–May 2014).

2.2 Elevation and volume change

For the measurement and understanding of decadal (between 2013–2016 and 2003–2005) elevation and volume change over GrIS, the difference map of AltiKa35 (March 2013–July 2016) and ICESat DEM (February 2003–June 2005)-derived elevation was generated. Furthermore, annual elevation changes were derived by differencing AltiKa cycles of the same month (so that surface condition will remain the same) between two consecutive years. Thereafter, elevation change map of cycle 11 (February–April 2014)–cycle 01 (March–April 2013), cycle 22 (March–April 2015)–cycle 11 (February–April 2014) and cycle 32 (March–April 2016)–cycle 22 (March–April 2015) were generated. Furthermore, outliers were removed using the inter-quartile range criterion. For obtaining the volume changes, the entire GrIS was divided into 19 basins, with basin Id ranging between 1.1 and 8.2 and is marked in figure 1. After multiplying elevation changes with the individual area of the basin, the results of volume change of the given basin was obtained.

3. Results and discussion

3.1 Validation of elevation and elevation changes of GrIS

The slope corrected elevation map derived over the GrIS using the SARAL/AltiKa dataset of cycle 01 (14 March–18 April 2013) is displayed in figure 2.

Figure 1. Drainage basin system of GrIS was adapted from Zwally *et al.* (2012).

Figure 2. Slope corrected elevation map of the GrIS was derived from SARAL/AltiKa cycle 01 (14 March–18 April 2013).

The elevation map indicated maximum elevation of GrIS at 3279 m with average elevation value of 2048 m. To quantify the accuracy of the slope uncorrected and corrected elevation map, we have compared the AltiKa-derived elevation with ICE-Sat DEM-derived elevations. It was observed that although the correlation and root mean square deviation (RMSD) remained similar with values around 0.95 and 267 m, respectively, for both slope

uncorrected and corrected elevation, bias value reduced to 48.31 m from 64.65 m. The large RMSD value can be attributed to the difference in period, along with interpolation methodology and undulating surface topography. As differences in slope corrected and uncorrected elevation is at a metre level and will not be visible in the case of the km-level elevation hence only slope corrected elevation map is displayed. In spite of large RMSD values,

we did further processing with elevations, as relative accuracy is high in comparison with absolute accuracy in the case of altimeter (highly precise instruments). Elevation change over the same surface will have a nearly similar/same error (Hurkmans *et al.* 2012).

Furthermore, to derive decadal elevation change, AltiKa35 elevations were corrected for penetration bias due to Ka band frequency. AltiKa-derived elevations (cycle 12, April–May 2014) when interpolated at ATM measured elevations of 5 April 2014, yielded bias of 0.38 m and 0.99 correlation for 2295 number of observation points over the GrIS region. This penetration bias is removed from the AltiKa35 elevation.

Annual elevation change results, over different regions of GrIS having different slope values were validated with ATM measurements which yielded a RMSD of 0.12 m for a relatively flat terrain (ROI-1) with a slope value of 0.19° and 1.79 m for ROI-2 with a slope value of 0.60° . This reveals the fact that the altimeter results are highly slope dependent, indeed RMSD increases with the increase in slope. Hence, we have limited our analysis to the region having slope values less than or equal to 1° which lead to 16% data reduction over the study region.

3.2 Impact of backscatter change on elevation change

The impact of backscatter change on elevation change derived using cycles 11, 22 and 32 has

been studied. Primarily, we have applied stringent data editing criteria ($-5.00 \text{ dB} \leq \Delta\sigma_0 \leq 5.00 \text{ dB}$, $-0.5 \text{ m} \leq \Delta H \leq 0.5 \text{ m}$, $\text{slope} \leq 0.1^\circ$) so that the ideal flat surface will be selected which can give maximum correlation (Davis and Ferguson 2004; Qiang *et al.* 2015) and measured correlation between elevation and backscatter change. This comparison yielded 0.12 (number of points = 12,675) (figure 3) and 0.03 (number of points = 12,026) correlation for difference of cycles 22, 11 and 32, 22, respectively.

From these results we can infer that selecting same season cycles for understanding elevation changes will not get affected by backscatter changes.

3.3 Elevation and volume changes of the GrIS

The difference map of AltiKa35 (March 2013–July 2016) elevation and ICESat DEM (February 2003–June 2005) elevations (figure 4) resulted in the negative volume change of $247 \text{ km}^3/\text{year}$ over the entire GrIS. During this period, irrespective of the elevation, all basins had negative elevation and volume changes. The maximum negative elevation change (0.36 m/year) was measured over basin 8.2 located in the north-west GrIS and the minimum negative elevation change (0.06 m/year) was over basin 7.1 located in the south-west GrIS. The coastal region of GrIS displays negative elevation change while the stable areas were located at the interior of GrIS (figure 4). The maximum negative volume change ($31 \text{ km}^3/\text{year}$) was measured over

Figure 3. Density plot showing correlation between elevation change (ΔH) and backscatter change ($\Delta\sigma_0$) between cycle 22 (Mar–Apr 2015) and cycle 11 (Feb–Apr 2014).

Figure 4. Elevation change map (after the removal of outlier) of GrIS for AltiKa (2013–2016)–ICESat (2003–2005).

basin 8.1 located in the north-west GrIS and the minimum negative volume change ($4 \text{ km}^3/\text{year}$) was measured over basin 1.4 located in the north-east GrIS. The maximum negative elevation and volume changes observed over basins 8.1 and 8.2 which were located in northwest region were in accordance with the results of Pritchard *et al.* (2010) who observed mass loss in the north, and the north-west region of GrIS using ICESat altimeter data for the period between 2003 and 2009. Furthermore, they attributed this mass loss to the glacier acceleration located in this area.

While deriving the annual elevation and volume change, results with cycle 01 were not discussed, as it has been observed that cycle 01 was not exactly in orbit of its past mission-Envisat and inclination manoeuvre have been applied to it (Chander *et al.* 2015; Remy *et al.* 2015). During 2014–2015 and 2015–2016 we observed a decrease in elevation and volume changes over the entire GrIS. Annual elevation changes during 2014–2015 (figure 5) indicate more than 50% of the decrease in elevation change which can be attributed to the very warm summer of July 2015, which prevailed along the north and north-western coast (nsidc.org/greenland.today/2015/11/2015-melt-season-in-review).

However, this extreme melting that occurred may be driven by the exceptional atmospheric condition characterised by new records in mean zonal winds and jet stream wave amplitude associated with the formation of arctic cut-off high

(Tedesco *et al.* 2016). The monotonic increase in the percentage of negative elevation change to 58% during 2015–2016 (figure 6) from 54% of 2014–2015, confirms the 2016 summer was a warmer summer than summer 2015 (nsidc.org/Greenland.today/2016/10/2016-melt-season-in-review).

Furthermore, we took three new ROIs as selected by Li and Davis (2008), to study the recent changes occurring over these ROIs. ROI-3 ($68.5\text{--}69.5^\circ\text{N}$, $46\text{--}50^\circ\text{W}$, the Jakobshavan glacier) indicates the negative change (2.07 m) over the Jakobshavan glacier during 2014–2015 and the positive change (0.31 m) during 2015–2016. During 2003–2005 to 2013–2016, ROI-3 displayed 1.39 m/year negative elevation change. ROI-4 ($68\text{--}71^\circ\text{N}$, $30\text{--}36^\circ\text{W}$, Kangerdlugssuaq) displayed 0.47 m positive elevation change during 2014–2015 and observed negligible negative elevation changes (0.03 m) during 2015–2016. During 2003–2005 to 2013–2016, negative elevation change with value 0.40 m/year was observed. ROI-5 ($64\text{--}67^\circ\text{N}$, $40\text{--}44^\circ\text{W}$, southeast glacier) displayed positive elevation change (0.90 m) during 2014–2015 and negative elevation change (0.14 m) during 2015–2016. During 2003–2005 to 2013–2016, negative elevation change with a value of 0.45 m/year was observed. This indicates that although annual change fluctuates between the positive and negative value, a decreasing trend was continued during the decadal changes over the selected ROIs as observed by Li and Davis (2008) during 1995–2006.

Figure 5. Elevation change map (after the removal of outlier) of GrIS for cycle 22 (Mar–Apr 2015)–cycle 11 (Feb–Apr 2014).

Figure 6. Elevation change map (after the removal of outlier) of GrIS for cycle 32 (Mar–Apr 2016)–cycle 22 (Mar–Apr 2015).

Moreover, annual basin-wise elevation changes (table 2) were derived. However, basins 8.1, 8.2, 1.1, 1.2 and 1.3 underwent negative elevation changes during both years 2014–2015 and 2015–2016 while positive elevation changes were measured in basin 3.3 located in the south-east. This indicates that the north and north-west region of GrIS where basins 8.1, 8.2, 1.1, 1.2 and 1.3 were located underwent a decrease in elevation change. Reported literature also indicates high thinning

rates in the north and north-west region of GrIS (Abdalati *et al.* 2001; Pritchard *et al.* 2010; Wang *et al.* 2018). Out of these five basins, four basins were located below 2000 m elevation while basin 3.3 was located at the highest elevation among all basins. This suggests that individual basin elevation may be the probable reason behind the observed changes, as it has been observed that the ice sheet above 2000 m has been almost exactly in balance for the past few decades (Thomas *et al.*

Table 2. Basin-wise inter-annual and intra-annual elevation and volume change estimates over GrIS.

Basin	Area $\times 10^6$ (km ²)	Elevation (m)	AltiKa-ICESat		Cycle 22–cycle 11		Cycle 32–cycle 22	
			Elevation change (m)	Volume change $\times 10^3$ (km ³)	Elevation change (m)	Volume change $\times 10^3$ (km ³)	Elevation change (m)	Volume change $\times 10^3$ (km ³)
1.1	0.126	1882.807	−2.293	−0.288	−0.216	−0.027	−0.558	−0.070
1.2	0.061	1932.392	−1.853	−0.113	−0.060	−0.004	−0.635	−0.039
1.3	0.044	1828.559	−1.922	−0.085	−0.009	0.000	−0.391	−0.017
1.4	0.017	1457.614	−3.191	−0.053	0.182	0.003	−0.107	−0.002
2.1	0.265	2276.417	−0.975	−0.258	0.188	0.050	−0.240	−0.064
2.2	0.050	2350.990	−1.088	−0.054	0.347	0.017	−0.238	−0.012
3.1	0.146	2735.779	−1.354	−0.198	0.510	0.074	−0.225	−0.033
3.2	0.036	1898.838	−2.253	−0.081	0.354	0.013	−0.573	−0.020
3.3	0.074	2817.283	−1.228	−0.091	0.263	0.019	0.065	0.005
4.1	0.066	2561.009	−1.096	−0.072	−0.065	−0.004	0.113	0.007
4.2	0.048	2220.534	−2.661	−0.128	0.455	0.022	−0.360	−0.017
4.3	0.035	2608.027	−2.349	−0.082	0.017	0.001	−0.407	−0.014
5.0	0.053	2262.384	−2.761	−0.146	−0.782	−0.041	−0.009	0.000
6.1	0.052	2338.403	−1.659	−0.087	−0.665	−0.035	0.276	0.014
6.2	0.140	2009.926	−1.964	−0.275	−0.354	−0.049	0.177	0.025
7.1	0.095	2558.784	−0.760	−0.073	−0.397	−0.038	0.093	0.009
7.2	0.129	2588.220	−1.288	−0.166	−0.457	−0.059	0.160	0.021
8.1	0.235	2361.490	−1.464	−0.344	−0.488	−0.115	−0.009	−0.002
8.2	0.032	1640.376	−4.008	−0.129	−0.419	−0.014	−0.032	−0.001

2001) and the low elevation shows widespread thinning (Abdalati *et al.* 2001). Thereafter, annual volume discharge by the entire GrIS during 2014–2015 and 2015–2016 was estimated to be 187 and 210 km³, respectively, indicating an enhanced decrease for the later period.

4. Conclusions

In the present study, we derived elevation, elevation and volume changes over GrIS using the SARAL/AltiKa 40 Hz GDR dataset of the ERM phase. We observed a maximum elevation with a value ~ 3279 m at the centre of the ice sheet and that continued to decrease towards the coast while average elevation was found to be 2048 m. Annual elevation changes derived over two different regions with a slope value 0.19° and 0.6° were compared with the ATM dataset yielded RMSD of 0.12 and 1.27 m, respectively. This reveals the fact that the altimeter results are highly slope-dependent, indeed RMSD increases with the increase in slope. Hence, data analysis was limited to regions having a slope value less than or equal to 1°.

Furthermore, the elevation difference map of AltiKa35 and ICESat DEM revealed a decreasing

rate of 247 km³/year by volume over the entire GrIS. Elevation changes obtained by differencing AltiKa cycles of the same month between two consecutive years, specifically during 2014–2015 and 2015–2016 were negative, over five basins located in the north and north-west, probably because of lower elevations, as four among them have an elevation less than 2000 m. Positive elevation change was observed over basin 3.3 located in the south-east region is located the highest among all. During 2014–2015 and 2015–2016 the entire GrIS contributed a volume loss of 187 and 210 km³, respectively, indicating an enhanced decrease for the later period.

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